Stratigraphy and paleogeographic significance of metamorphic rocks in the Shadow Mountains, western Mojave Desert, California

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ABSTRACT

Stratigraphic correlations presented here for the ductilely deformed and metamorphosed rocks exposed in the Shadow Mountains indicate that they formed on the North American continental margin and are not exotic or significantly displaced from their site of origin. These strata represent a depositional history that spans Late Proterozoic and Paleozoic passive margin development, late Paleozoic transitional passive to active plate margin tectonics, and late Paleozoic–early Mesozoic establishment of a convergent margin along the western edge of the North American craton.

The stratigraphic sequence in the Shadow Mountains is represented by a basal siliceous and calcareous section that is correlated with upper Proterozoic and Cambrian miogeoclinal strata. These strata are overlain unconformably(?) by a calcareous sequence correlated with rocks of Pennsylvanian and Permian age of borderland affinity. The uppermost sequence comprises hornfels and calcareous rocks that rest unconformably across older strata and are correlated with Permian and Triassic rocks that likely record uplift and erosion of a magmatic arc.

These stratigraphic correlations have several important paleogeographic and tectonic implications for southwestern North America. First, passive margin, Late Proterozoic and early Paleozoic miogeoclinal facies extend as far west in the Mojave Desert as the Shadow Mountains. Second, the presence of Pennsylvanian and Permian borderland basin sediments suggests that the area was affected by late Paleozoic tectonics, probably associated with the transition from a passive to an active plate-margin setting. Third, inferred Late Permian–Triassic rocks record the onset of convergent-margin tectonism and magmatism in this region.

INTRODUCTION

The Shadow Mountains comprise the largest exposure of metasedimentary rock in the central and western Mojave Desert. Previous workers interpreted these rocks to range from Late Proterozoic to Mesozoic in age (Bowen, 1954; Troxel and Gunderson, 1970; Poole, 1974; Burchfiel and Davis, 1975, 1981; Miller and Cameron, 1982; Brown, 1983). Despite this earlier work, no consensus exists for either the age or correlation of these strata. A better understanding of these rocks is necessary to place them within the frame-work of the Late Proterozoic to early Mesozoic tectonic development of the southwestern Cordillera. In addition, the affinity of the Shadow Mountain strata is critical to evaluating models for the Mesozoic and Paleozoic paleogeography of the southwestern Cordillera, which call for major tectonic boundaries to be present in the area (e.g., Walker, 1988; Stone and Stevens, 1988; Lahren et al., 1990).

The significance of the Shadow Mountains stratigraphy for understanding the regional geology can best be appreciated by reviewing our current knowledge of the Proterozoic and Paleozoic tectonic history of this region. The Late Proterozoic and Paleozoic western margin of North America, as defined by cratonal, miogeoclinal, and eugeoclinal facies patterns, generally trends northeast (Fig. 1; Burchfiel and Davis, 1975, 1981). Paleozoic miogeoclinal and cratonal facies project southwestward from southern Nevada and southeastern California into the Mojave Desert, where they intersect the northwest-trending Mesozoic plutonic arc at a high angle (Hamilton and Myers, 1966; Burchfiel and Davis, 1975; Stewart and Poole, 1975). Rocks of eugeoclinal character occur from the El Paso Mountains to Lane Mountain (Fig. 2; Poole, 1974; Carr et al., 1981; Wust, 1981; Miller and Sutter, 1982). A notable observation from these studies is that Paleozoic eugeoclinal rocks are juxtaposed against like-age transitional miogeoclinal-cratonal rocks with the absence of intervening thick miogeoclinal sequences and middle Paleozoic clastic rocks related to the Antler orogeny (Walker, 1988; Martin and Walker, 1991, 1992). Both the extent and tectonic significance of these eugeoclinal rocks in the Mojave Desert have been discussed and questioned by many workers. Davis et al. (1978) and Burchfiel and Davis (1981) attributed the presence of eugeoclinal rocks to strike-slip faulting during Permo-Triassic truncation of the continental margin. Alternatively, Poole (1974) considered the eugeoclinal rocks to be essentially in place, but telescoped with miogeoclinal rocks by late Paleozoic and Mesozoic thrusting. Various combinations of these interpretations have been utilized by later workers (e.g., Stone and Stevens, 1988; Walker, 1988; Snow, 1992).

The main obstacles to interpreting the nature of this juxtaposition have been (1) the difficulty in determining the age and paleogeographic affinity of many exposures across the central Mojave Desert due to degree of metamorphism and deformation associated with development of the Mesozoic Cordilleran arc and (2) the lack of exposures of contacts between known eugeoclinal and miogeoclinal-cratonal rocks within the same area.

Since late Paleozoic–Early Triassic time, the central and western Mojave has been along an active plate margin. As a result of deformation and magmatism associated with this tectonic setting, pre-Mesozoic paleogeographic and tectonic trends present within

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and west of the central Mojave Desert are not preserved or are strongly obscured. To determine what these trends were, it is paramount that we understand the age and paleotectonic affinity of the metasedimentary rocks exposed along the western part of this margin such as those that crop out in the Shadow Mountains. The aim of this paper, therefore, is to present stratigraphic descriptions and correlations from the Shadow Mountains in order to interpret the Late Proterozoic to Mesozoic paleogeographic development of the southwestern Cordillera.

STRUCTURE AND METAMORPHISM

Sedimentary rocks in the Shadow Mountains are intensely deformed and record metamorphism within the amphibolite facies. Because deformation and metamorphism limit the stratigraphic correlations made in this paper, a brief discussion of these features is given here. Original sedimentary layering is transposed by foliation (formed during deformation event D1) that has been folded by at least two subsequent phases of coaxial north-south–trending, dominantly west-vergent ductile folds. The first folding event is represented by a kilometer-scale overturned, nearly recumbent antcline (D2) in the northern Shadow Mountains (Silver Peak anticline, Fig. 3). At least one subsequent phase of folds (D3) overprints D2 folds. The foliation and folds are associated with amphibolite-facies metamorphism (Martin, 1992). Because of deformation, a significant amount of shearing has occurred along lithologic contacts. This is particularly pronounced in the northern Shadow Mountains around the limbs of the map-scale Silver Peak and Shadow Valley anticlines (Fig. 3); lithologic units are discontinuous around these folds. Although the discontinuous nature of these units is considered related to attenuation and ductile shearing associated with deformation, factors such as nondeposition and/or erosion cannot be precluded (see stratigraphic descriptions, Appendix, and following discussion).

Development of the D1 foliation, D2 and D3 folds, and associated metamorphism is older than 148 Ma, the age of the postkinematic, bimodal gabbro/granite complex in the Shadow Mountains (U-Pb zircon; Martin, 1992). Intrusion of the igneous complex imparted a pervasive, static greenschist- to lowest amphibolite-facies retrograde metamorphism on the older amphibolite-facies dynamothermal metamorphism. U-Pb geochronology on both metamorphic and detrital zircons that have experienced high-temperature Pb loss suggests that dynamothermal metamorphism occurred at 165 ± 11 Ma (Martin, 1992). Based on the structural and metamorphic relations and the U-Pb geochronology, D1 foliation, folding, and metamorphism are interpreted to have formed during a period of progressive contractile deformation; however, older deformation cannot be precluded. Based on a similar style of ductile contractile deformation and permissive timing constraints, this deformation is tentatively correlated with Middle to Late Jurassic contractile deformation identified in numerous ranges to the northeast, between the Shadow Mountains and the Garlock fault (Walker et al., 1990; Martin, 1992; Boettcher and Walker, 1993).

STRATIGRAPHIC CORRELATION

Metasedimentary rocks in the Shadow Mountains can be divided into three assemblages, which can be further subdivided into two or more map units (Fig. 4). The areal distribution of these units is shown in Figure 3. In the following section we present possible
correlations for each of these stratigraphic assemblages. A detailed description of these units and their thicknesses is given in the Appendix.

Assemblage 1

Assemblage 1 is subdivided into five map units (Fig. 3): unit $C_1$ consists of quartz-biotite schist containing minor quartzite; unit $C_2$, pink to white quartzite and meta-arkose with minor pebble to cobble conglomerate and quartz-biotite schist; unit $C_3$, thin- to medium-layered siliceous calcite marble, calc-silicate rocks, and metasiltite with minor dolomite marble and quartzite; unit $C_4$, massive dolomite marble and, as mapped, structurally interleaved layers of siliceous and buff colored calcite marble; and unit $C_5$, white to tan dolomite marble intercalated with and overlain by compositionally variable quartzites, meta-arkose, and quartz-biotite schist.

Units of assemblage 1 are correlated with the upper Proterozoic to Cambrian miogeocline sequence including the Wood Canyon Formation, Zabriskie Quartzite, Carrara Formation, and Bonanza King Formation. This correlation is based on overall stratigraphic succession and comparison with a nearly identical sequence exposed at Quartzite Mountain to the southeast (Fig. 2), which has been identified as upper Proterozoic and Cambrian (Stewart and Poole, 1975; Miller, 1981), as well as on numerous other studies (Kiser, 1981; Brown, 1983, 1991; Glazner et al., 1988; Martin and Walker, 1991; Boettcher and Walker, 1993). We correlate quartz-biotite schists of unit $C_1$ with the Wood Canyon Formation and the overlying quartzite of unit $C_2$ with the Zabriskie Quartzite (Fig. 3). Detrital zircon U-Pb geochronology and whole-rock Nd and Sr isotopic studies on a meta-arkose bed from unit $C_2$ support a 1.7 Ga provenance age ($\epsilon_{Nd} = -18, {^{87}Sr/^{86}Sr = 0.7376}$), in agreement with 1.35 and 1.8 Ga lower to middle Proterozoic basement in the area (Martin and Walker, 1992) and the Zabriskie Quartzite correlation. The calcareous rocks of unit $C_3$ are correlated with the Cambrian Carrara Formation.

Brown (1983) correlated rocks of unit $C_4$ with the Bonanza King Formation (Fig. 3), a correlation with which we concur. Although not separated during mapping, siliceous and buff-colored marble layers transposed within and above the Bonanza King Formation may represent the Dunderberg Shale and Nopah Formation, and perhaps even Devonian and Mississippian strata. A thin middle Paleozoic sequence lacking Ordovician and Silurian strata in the Shadow Mountains would be consistent with the interpretation presented by Martin and Walker (1991, 1992, 1993), in which transitional miogeocline-cratonial rocks continued west-southwestward through the central Mojave Desert. However, transposition and folding in the Shadow Mountains precludes unequivocal stratigraphic determinations for these rocks.

Figure 2. Map of pre-Tertiary rocks in the central and western Mojave Desert. Mojave Valley fault (MVF) is marked by a dashed line; offset by numerous northwest-trending, late Cenozoic strike-slip faults is not shown. Map modified from Martin et al. (1993, Fig. 1).
Assignment of unit pCu is problematic (Fig. 3). This unit is lithologically dissimilar to the overlying assemblages 2 and 3, and to other middle Paleozoic to Mesozoic sequences in the Mojave Desert. The gross succession of white to tan dolomite marble intercalated with and overlain by compositionally variable quartzites, meta-arkose, and quartz-biotite schist is most similar to upper Proterozoic to Cambrian rocks of the Johnnie Formation, Stirling Quartzite, and the lower to middle Wood Canyon Formation.
This correlation is favored here and is consistent with this unit occupying the lowest apparent stratigraphic position in the northern Shadow Mountains, although exact correlation is precluded by extreme deformation.

**Assemblage 2**

The rocks of assemblage 2 are generally well exposed and have been divided into four map units: Unit PP₁ consists of thin-to-medium-layered calcite marble that contains numerous sedimentary structures indicative of mass-flow deposits; unit PP₂, orange to brown schistose feldspathic quartzite; unit PP₃, white to tan massive dolomite to calcite marble; and unit PP₄, a heterogeneous mix of various types of calcareous marbles, locally containing sedimentary structures indicative of mass-flow deposits.

Two poorly preserved brachiopods suggestive of a Pennsylvanian age have been the only known fossils collected from the Shadow Mountains. These fossils were collected by Bowen (1954) from the southeastern part of the area (W½ sec. 25, T. 7 N., R. 6 W.; Dibblee, 1967) and appear to have been taken from unit PP₁₋₃ of this assemblage (Fig. 3). Thirty-four samples were collected during this study from various marble beds throughout assemblages 2 and 3 in an effort to recover conodonts. All samples were barren (B. R. Wardlaw, 1992, personal commun.), probably due to the high degree of metamorphic recrystallization.

Miller and Cameron (1982) noted similarities of the stratigraphy in the southern Shadow Mountains (assemblage 2) and upper Paleozoic rocks in the Lane Mountain–Goldstone area, El Paso Mountains, and the Inyo Mountains (Fig. 1). On the basis of our new interpretations (Appendix), such as the likely existence of rocks of mass-flow origin in assemblage 2 and the limited fossil control (see above), we concur with the assessment made by Miller and Cameron (1982). Basinal calcareous turbidites in the Inyo Mountains and Darwin Hills (Fig. 5) are of Pennsylvanian and Permian age (Stevens and Stone, 1988). These strata are said to have “borderland” affinity because they were deposited in isolated, rapidly subsiding and shoaling basins interpreted to have been related to left-lateral strike-slip and thrust faults associated with truncation of the southwest margin of North America during Pennsylvanian and Permian time (Stone and Stevens, 1984, 1988; Stevens and Stone, 1988; Walker, 1988).

Besides lithologic similarities to rocks of borderland affinity, the regional stratigraphic occurrence of these rocks is also important for our correlation of assemblage 2 rocks. Rocks of borderland affinity rest conformably on Lower Pennsylvanian and Mississippian miogeoclinal rocks in the southern Inyo Mountains and Darwin Hills (Fig. 5; Stone and Stevens, 1988). In the Soda Mountains, Lower Permian borderland sedimentary rocks rest conformably on transitional miogeoclinal-cratonal Pennsylvanian and Mississippian rocks (Figs. 2 and 5; Walker and Wardlaw, 1989). Lower Permian
Figure 5. Correlation diagram incorporating composite sections from the El Paso Mountains (Carr et al., 1984), Quartzite Mountain area (Miller, 1981), Soda Mountains (Walker and Wardlaw, 1989), southern Inyo Mountains and Darwin Plateau area (Stone and Stevens, 1984). Figure modified from Walker (1988).
turbidites in these areas are unconformably overlain by Upper Permian to Lower Triassic strata (Fig. 5). In the El Paso Mountains (Figs. 2 and 5), Ordovician and Devonian eugeoclinal rocks (Antler belt) and Mississippian Antler foreland rocks are deformed and overlain by Pennsylvanian and Permian rocks (Poole, 1974; Carr et al., 1981, 1984). Pennsylvanian strata in this area comprise shallow-water carbonates that grade upward into Pennsylvanian and Lower Permian turbidites (Carr et al., 1984). Lower Permian rocks contain Guadalupian to Wolfcampian (?) mass-flow and olistostrome deposits. These rocks are overlain by volcanogenic debriscontain Guadalupian to Wolfcampian (?) mass-flow and olistostrome deposits. These rocks are overlain by volcanogenic debris that contains radiogenic component.

Although the mechanism by which eugeoclinal and transitional miogeoclinal-cratonal strata are juxtaposed is debated (Stevens and Stone, 1988; Walker, 1988; Snow, 1992), it is clear that Pennsylvanian and Permian borderland sedimentation was associated with both allochthonous and autochthonous strata (Stevens and Stone, 1988; Walker, 1988). Therefore, aside from their overall lithologic similarity to borderland rocks, the association of inferred Pennsylvanian and Permian borderland strata in the Shadow Moun-

tains with miogeoclinal and/or transitional miogeoclinal-cratonal strata is consistent with regional stratigraphic relationships.

**Assemblage 3**

Assemblage 3 comprises two units: PT1 and PT2 (Fig. 4). Unit PT1 consists of massive hornfelses containing lesser amounts of quartzite, schistose feldspathic quartzite, calc-silicate rocks, calcite marble, and talc-chlorite schist. Unit PT1 is gradational into unit PT2 (Fig. 4), which is composed of calcite marble, calc-silicate rocks, and lesser amounts of siliceous hornfels, quartz-biotite schist, and quartzite. Unit PT2 protolith was made up predominantly of immature fine- to coarse-grained clastic rocks with minor amounts of marine carbonate rocks. Although equivocal and sparse, rocks of volcanic origin may also be present. Unit PT2 is dominantly marine in origin. Assemblage 3 is interpreted to rest unconformably on assemblage 2 and possibly also on assemblage 1 (Appendix).

Our interpretation is that units PT1 and PT2 of assemblage 3 (Fig. 3) correlate with Upper Permian rocks in the El Paso Mountains and Lower Triassic rocks in the central Mojave Desert. To appreciate this correlation we must discuss the stratigraphic relations in the El Paso Mountains. In addition, because relations in the Shadow Mountains only bracket these rocks to be between Pennsylvanian-Permian and Middle Jurassic in age, we present new age data for El Paso Mountains strata that helps establish the age of assemblage 3 rocks.

In the central and eastern El Paso Mountains, the succession from Pennsylvanian and Lower Permian borderland rocks gives way upward to Upper Permian volcanic rocks (Carr et al., 1984); this transition is considered the earliest indication of the Sierran arc (Burchfiel and Davis, 1981; Miller and Sutter, 1982; Carr et al., 1984; Walker, 1988; Burchfiel et al., 1993). In the Goler Gulch area this transition is marked by an abrupt change from borderland carbonate rocks to a thick succession of andesite flows. In the Iron Canyon area, west of Goler Gulch, the contact between borderland sediments and andesite flows is gradational across a succession of volcanic and volcanogenic rocks with intercalated carbonates (Fig. 5; Carr et al., 1984).

A similar succession exposed in the western El Paso Mountains,
the Bond Buyer Sequence, contains, in ascending order, andesitic flows, clastic, and calc-silicate rocks that are intruded locally by Early to Middle Triassic granodiorites (227–239 Ma, Cox and Morton, 1986, written commun.; 246 Ma, Miller et al., 1993, unpub. data; Walker, 1988). Unfortunately, the correlation of the Bond Buyer Sequence with the more intact upper Paleozoic succession in the central El Paso Mountains is uncertain (Carr et al., 1984; Walker, 1988). The clastic and calc-silicate rocks in the Bond Buyer Sequence have been correlated with rocks of similar composition in the Goldstone–Lane Mountain area, which have been correlated with the Lower Triassic Fairview Valley Formation, exposed in the Quartzite Mountain area (Fig. 5; Burchfiel et al., 1980; Walker et al., 1984; Walker, 1988). Alternatively, the Bond Buyer has been considered equivalent to the andesites described above (Carr et al., 1984).

To determine the age of the post–Lower Permian stratigraphic successions in the El Paso Mountains, two volcanic samples were collected for U-Pb zircon geochronology: Andesite-A and Andesite-D (Table 1, Fig. 6). Andesite-D was collected from the base of the volcanic section in Goler Gulch and yields an early Late Permian age of 262 ± 2 Ma (Model 1 of Ludwig, 1990). This age is consistent with the age of underlying Lower Permian turbidites (Carr et al., 1984) and establishes a minimum age for the volcanic sequence exposed in the central and eastern El Paso Mountains. Andesite-A was collected from the Bond Buyer Sequence exposed in the western El Paso Mountains and yields an Early Permian age of 281 ± 8 Ma (Fig. 6; Model 1 of Ludwig, 1990, and forced through 0 Ma to average the Pb/Pb ages).

This latter age implies that the Bond Buyer Sequence represents yet another transitional section from the borderland to the arc succession. This age, combined with ages from nearby plutons, makes it unlikely that the Bond Buyer correlates with the Fairview Valley Formation or any other Lower Triassic rocks in the Mojave Desert region (e.g., Walker, 1988); rather, it is entirely an upper Paleozoic succession (e.g., as interpreted by Carr et al., 1984).

We consider the transition from Pennsylvanian and Lower Permian borderland rocks to Upper Permian arc-derived volcanic and sedimentary rocks with intercalated marine carbonates in the central and eastern El Paso Mountains to be grossly similar to the transition from assemblage 2 calcareous mass-flow deposits to assemblage 3 siliceous hornfels and intercalated marbles (unit PT1) in the Shadow Mountains (Figs. 4 and 5). Although the presence of volcanic rocks in assemblage 3 is equivocal, it is permissible for the siliceous hornfels of assemblage 3 to be arc-derived clastic rocks. The angular discordance between assemblages 2 and 3 described here is not observed across the El Paso Mountain transition. However, considering the differences between the two sections described in the central and eastern El Paso Mountains (discussed above), we do not believe that exact lithologic or stratigraphic similarities should be expected between ranges.

The following paleogeographic setting is proposed for arc-derived units of assemblage 3. Meta-clastic rocks of unit PT2 are correlated with Upper Permian volcanogenic and volcanic rocks in the central El Paso Mountains (Fig. 5) and probably also in the Goldstone and Lane Mountain area. This unit represents Upper Permian clastic rocks derived from the proto-Sierran magmatic arc and deposited in a marine or near-marine environment. Local and/or regional uplift associated with arc development and associated deformation accounts for the angular unconformity between assemblage 3 and older strata.

Although a reasonable maximum age for unit PT2 has not been established, it must be older than 165 Ma (age of metamorphism in the Shadow Mountains, Martin, 1992). Because unit PT2 appears lithologically gradational with unit PT1 (Appendix), we infer that it was deposited relatively close in time to unit PT1. Based on this inference, marbles and calc-silicate rocks of unit PT2 suggest either the diversion of arc sediments (represented by unit PT1) or the erosion of the arc to near sea level by this time. In this context, unit PT2 is tentatively correlated with the Lower Triassic overlap sequence, the Fairview Valley Formation, exposed in the Quartzite Mountain area, and correlative rocks in the Lane Mountain–Goldstone area and Soda Mountains (Fig. 5; Burchfiel et al., 1980; Miller, 1981; Walker et al., 1984; Walker, 1988).

Figure 6. U-Pb concordia diagram for samples Andesite-A and Andesite-D. Sample locations and analytic data are presented in Table 1.
DISCUSSION

Figure 7 represents a series of time slices that show schematically the paleogeographic and tectonic setting in the region of the central and western Mojave Desert from Late Proterozoic to early Mesozoic time. The paleogeography presented in Figure 7 incorporates many aspects of earlier models that call on late Paleozoic–Early Triassic truncation of the southwestern Cordillera to explain (1) the angular discordance between the Late Proterozoic–Paleozoic facies trends and the trend of later Mesozoic features (Burchfiel and Davis, 1975; Walker, 1988), (2) the juxtaposition of eugeoclinal rocks and Antler foreland deposits with miogeocinal-cratonal rocks in the Mojave Desert (Burchfiel and Davis, 1975; Poole and Christiansen, 1980), and (3) the tectonic setting during the deposition of Pennsylvanian-Permian borderland rocks defined by Stone and Stevens (1984) and Stevens and Stone (1988).

Late Proterozoic to Mississippian paleogeography in the Mojave Desert region was characterized by a northwest-facing passive margin (Fig. 7A). On the basis of regional stratigraphic data discussed by Martin and Walker (1991, 1992, 1993), we extend the miogeocinal-cratonal hinge line (defined as those sections where lower to middle Paleozoic sequences are relatively thin or missing,
but upper Precambrian to Cambrian sections are relatively well developed) through the Mojave Desert at least as far west as the Shadow Mountains. The Mississippian Antler orogenic belt parallels Paleozoic facies trends through central Nevada into central California. This interpretation implies that these facies and tectonic trends continued to the southwest.

Pennsylvanian and Early Permian time was a period of transition from passive to active plate-margin setting along the southwestern Cordillera (Figs. 7B and 7C). Following Walker (1988), during this time period Antler belt rocks moved southward along a south-southeast–striking, sinistral strike-slip fault or fault system, truncating older northeast-striking passive margin facies patterns. Transpressional and/or transtensional structures associated with this tectonic setting most likely resulted in local uplift and basin formation (e.g., Stone and Stevens, 1988; Stevens and Stone, 1988). Pennsylvanian and Permian borderland sediments (diagonal ruling in Figs. 7B and 7C) were deposited in basins associated with both allochthonous Antler belt rocks (El Paso Mountains and areas in the north-central Mojave Desert) and autochthonous miogeoclinal and transitional miogeoclinal-cratonal rocks (Inyo Mountains, Darwin Hills, Soda Mountains, and Shadow Mountains). In the central and western Mojave Desert, local uplift and erosion associated with this tectonism may have caused erosion of preexisting Paleozoic strata, which may in part explain missing section (not ignoring structural modification) between assemblages 1 and 2 in the Shadow Mountains. Observations and interpretations presented here for the Shadow Mountains imply that the central and western portions of the Mojave Desert were affected by late Paleozoic basin formation and deformation. Considering older facies trends, this suggests encroachment of these modifications into more interior parts of the Cordillera and implies that the major structure(s) that truncates the margin must lie outboard of the Shadow Mountains.

By Late Permian time (Fig. 7D), allochthonous Antler belt rocks docked with Paleozoic transitional miogeoclinal cratonal rocks in the Mojave Desert, in part explaining the absence of a fully developed miogeoclinal facies in the central and western Mojave Desert. Proximal to a marine environment, magmatic arc activity and associated contractile deformation began in the Mojave Desert area by Late Permian time with the onset of convergent margin activity (diagonal ruling, Fig. 7D). Rocks belonging to assemblage 3 and the apparent angular discordance between assemblages 2 and 3 in the Shadow Mountains are interpreted to record the uplift and erosion of this arc, probably in a marine or near-marine environment.

Snow (1992) proposed an alternative model that calls on the Permian (?) Last Chance Thrust System to explain late Paleozoic structural and stratigraphic relations in the region west of Death
Valley and in the Mojave Desert. Although our observations presented here do not contradict this model, the truncation model relies on fewer inferred structural projections (southward bend of the Antler belt and southern continuation of the Last Chance Thrust System into the Mojave Desert) and changes in facies trends (miogeoclinal becoming narrower southward into the Mojave Desert). In addition, east- or southeast vergent structures that are present in the Mojave Desert are demonstrably Middle to Late Jurassic (Walker et al., 1990; Boettcher and Walker, 1993; Glazner et al., 1994) and not late Paleozoic (Snow, 1992) in age. Finally, U-Pb geochronology and isotopic data from Permian to Mesozoic plutonic rocks suggest that eugeoclinal rocks in the Mojave Desert area cannot have been thrust eastward over Proterozoic continental crust (as implied by Snow, 1992) until between Middle Triassic and Middle Jurassic time (Miller et al., 1993, unpub. data).

SUMMARY

The correlations presented here for the Shadow Mountain stratigraphy help elucidate the Late Proterozoic to early Mesozoic paleogeography of the central and western Mojave Desert. Assemblage 1 rocks are correlated with Late Proterozoic–Cambrian miogeoclinal rocks indicative of passive margin development. Assemblage 2 is correlated with Pennsylvanian to Early Permian borderland sedimentary rocks. This correlation implies that deformation associated with the deposition of these rocks had encroached on interior parts of the Cordillera. Siliceous hornfels, marbles, and calc-silicate rocks of assemblage 3 are correlated with Late Permian and Triassic sequences elsewhere in the region that represent the development of a volcano-magmatic arc in or near a marine environment. This sequence represents the onset of convergent margin tectonics in this part of the Cordillera. These correlations indicate that autochthonous continental North America extends westward from the eastern Mojave Desert to the Shadow Mountains.

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APPENDIX: STRATIGRAPHIC DESCRIPTION

Note: All thicknesses reported for the units described in the Shadow Mountains are tectonic.

Assemblage 1

Description. Assemblage 1 is exposed in a nearly recumbent anticline in the northern Shadow Mountains (Fig. 3) and is divided into five map units. Unit CE1 (Fig. 3) occurs in the core of the anticline at Silver Peak. It is 20–30 m thick and composed of coarse- to medium-grained, quartz-biotite schist that locally contains thinly layered, medium-grained quartzite. Unit CE1 is conformably overlain by unit CE2 (Fig. 3), which consists of 30–40 m of coarse- to fine-grained, pink to white quartzite and meta-arkose. This unit contains discontinuous layers of granite-pebble to cobble conglomerate, and minor amounts of quartz-biotite schist. Possible relict Skolithos tubes were observed in one quartzite bed. Unit CE3 (Fig. 3) rests conformably on unit CE2 and consists of thin- to medium-layered, siliceous calcite marble, calc-silicate rocks, and metasiltite with minor dolomite marble. The upper portion of this calcareous sequence contains thin, laterally extensive, fine- to medium-grained green and pink quartzite and quartz-biotite schist horizons. The thickness of unit CE3 is somewhat variable due to ductile attenuation and shearing related to folding, but the maximum thickness is ~120 m. Unit CE4 (Fig. 3) consists of massive dolomite marble ~120–150 m thick. Structurally interleaved within the dolomite marble are laterally discontinuous layers of siliceous and buff-colored calcite marble that are included within unit CE3 because of their transposition with the dolomite marble. The contact between unit CE3 and unit CE4 has been strongly modified due to the competence contrast between these two units and attenuation related to ductile folding. As a result, unit CE4 is only present along the large rootless parasitic fold (Shadow Valley anticline) in the uplift limb of the Silver Peak anticline (Fig. 3; Martin, 1992).

In addition to this sequence of four units, a sequence of rocks containing, in ascending structural order, white to tan dolomite marble with thin interlayered quartzite and meta-arkose beds, quartz-biotite-staurolite schist and dark-green to blue, medium- to coarse-grained (locally pebbly) micaeous (locally calcareous) and amphibole-bearing quartzite, and pink to white quartzite and quartz-pebble quartzite is exposed in the east-west-trending Silver Peak area (unit pCEu, Fig. 3). This sequence of rocks forms the core of the overturned anticline in an area that is complicated by folding and structural attenuation of these rocks. This group of rocks has a combined thickness of ~300–400 m and is designated unit pCEu.

Assemblage 2

Description. Rocks assigned to assemblage 2 consist of calcite marble, conglomeratic marble, siliceous calcite marble, calc-silicate, dolomite marble, quartzite, and meta-laminate. Assemblage 2 (units PP1–4, Fig. 3) is widely exposed in the southern Shadow Mountains. In the northern Shadow Mountains, a thin sequence assigned to assemblage 2 lies between Cambrian Carrara Formation and assemblage 3 rocks and in high-angle fault contact against Bonanza King Dolomite (Fig. 3). Rocks in the southern Shadow Mountains are at lower metamorphic grade and less deformed than those in the northern Shadow Mountains (although the base of assemblage 2 is not exposed). For this reason, the following description is based mainly on the southern exposures where assemblage 2 can be divided into four units (Fig. 4) and units PP1–4, of Fig. 3).

Unit PP1 (Fig. 3) is composed of thin- to medium-layered, dark blue to gray, resistant calcite marble. Within this unit are 0.5- to 2.0-m-thick layers of angular carbonate-pebble to cobble-collongenrate overlain by metagrainstones and finer-grained, laminated marbles. These fining-upward variations in grain size generally occur within the same bed. Clasts within conglomerate beds are <0.5 m in diameter and are composed of gray calcite marble (Fig. 8). Metagrainstone beds commonly preserve apparent fining upward trends (Fig. 9). Although recrystallized, these grainstone horizons locally contain possible relict fusulinids and/or crinoid ossicles. Locally, finer-grained, thinly laminated siliceous marbles rest on grainstone beds (Fig. 10). Unit PP1 also contains 2- to 10-m-thick layers of alternating thinly layered to platy, siliceous calcite marble and calcite marble as well as local, pure white quartzite or calcareous quartzite layers (<0.5 m thick). Several white calcite marble beds contain rounded, medium-grained quartz grains. Another distinctive, but rare rock type is brown to dark green, quartz-mica-amphibole-andalusite hornfels and schist similar to some rocks of assemblage 3 (described below). The mineralogy suggests the protolith was an Al-rich pelite. Unit PP1 has a minimum thickness of 500 m; the base is not exposed.

Unit PP2 (Fig. 3) rests on unit PP1, and consists of orange to brown, thin- to medium-layered, medium-grained, schistose feldspathic quartzite; this rock is gradational over a few meters from siliceous calcite marble to metacalcareous siltstone to meta-arkose. The upper 1–2 m of unit PP2 consists of interlayered metasiltite and calcareous marble. Unit PP2 is overlain by 20–30 m of featureless, white to tan, massive dolomite to calcite marble of unit PP3 (Figs. 3 and 4).
Unit PP3 (Fig. 3), which overlies unit PP2, and has a minimum thickness of ~750 m, is lithologically heterogeneous. The contact between units PP3 and PP2 is relatively abrupt and is marked by the first appearance of siliceous sandstone or eolian sand origin can be discounted. The vertical transition from unit PP4 to unit PP3, assemblage 2 (unit PP2), Fig. 3) rests on rocks correlated with Cambrian Carrara Formation (unit Cc, Figs. 3 and 4). Strata above and below this contact are discordant on the upright limb of the anticline; around the hinge, strata inferred to belong to lower assemblage 2 are discontinuous and are missing on the overturned limb of the fold where assemblage 3 rests against Carrara Formation (unit Cc, Fig. 3). Elsewhere in the northern Shadow Mountains, carbonate rocks of assemblage 2 lie beneath assemblage 3 and in high-angle fault contact with Bonanza King Formation (unit Cc, Fig. 3). In part, the apparent absence and discontinuous nature of assemblage 2 units along the folds in the northern Shadow Mountains is due to attenuation and ductile shearing related to deformation, although nondeposition and/or erosion of these units prior to deformation cannot be precluded.

Interpretation. The presence of fine-grained carbonate beds (conglomerate to laminated marble) in unit PP2 suggests a mass-flow origin. The periodic influx of siliceous material associated with unit PP2 may suggest a complex depositional setting. The presence of sparse siliceous horizons similar in mineralogy to that in the stratigraphically higher assemblage 3 may suggest a magmatic-volcanic arc source for these rocks (discussed below).

The absence of sedimentary structures in the schistose feldspathic quartzite of unit PP2 prevents exact determination of its depositional environment. However, the abundance of metamorphic muscovite and andalusite suggests that Al-rich clays are present in the protolith and that a winnowed beach sand or eolian sand origin can be discounted. The vertical gradation from marble to quartzite and then back into marble suggests a gradual transition to a subaerially-derived regime and back to carbonate deposition within a marine environment.

We interpret the massive marble of unit PP2 as indicative of a carbonate-platform depositional environment. In unit PP2, the cross-bedded nature of the marbles and the sharp truncation of these structures by siliceous beds are concordant on the upright limb of the anticline; below this contact are concordant on the upright limb of the anticline; whereas above this contact are concordant on the overturned limb of the fold where assemblage 3 rests against Carrara Formation (unit Cc, Fig. 3). Elsewhere in the northern Shadow Mountains, carbonate rocks of assemblage 2 lie beneath assemblage 3 and in high-angle fault contact with Bonanza King Formation (unit Cc, Fig. 3). In part, the apparent absence and discontinuous nature of assemblage 2 units along the folds in the northern Shadow Mountains is due to attenuation and ductile shearing related to deformation, although nondeposition and/or erosion of these units prior to deformation cannot be conclusively demonstrated. Based on the available data, assemblage 3 is interpreted to rest unconformably on assemblage 2 and possibly also assemblage 1 rocks.

Because the top of assemblage 3 is not observed, the minimum exposed structural thickness is 2000 m in the northwest Shadow Mountains. (Fig. 3).

References Cited


Bown, H., 1953, Possible Cambrian monocline strata in the Shadow Mountains, western Mojave Desert, California: Geological Society of America Abstracts with Programs, v. 5, p. 413.


